LITHOSPHERIC MODELLING OF VENUS

W. B. Banerdt and M. P. Golombek, Jet Propulsion Laboratory, California Institute of Technology, Pasadena, CA 91109

Introduction: Simple rheological models, such as an elastic layer over a viscous or inviscid half space, have been used to investigate the structure and behavior of planetary lithospheres. A more realistic rheology, where the failure stress at a given depth is the weaker of the frictional strength (determined from Byerlee's law) or the ductile strength (which is highly material dependent), can be represented by a lithospheric strength envelope [1,2]. In this abstract we investigate some of the implications of this type of rheology for modelling tectonic structures found on Venus.

Rift Formation: Because the ductile strength of mantle materials is much greater than that of the crust, the mechanical structure of the lithosphere is critically dependent on the crustal thickness (e.g. [2]). For a crust on Venus thicker than about 30 km, there is no appreciable strength in the mantle (assuming a surface temperature of 740°C, a thermal gradient of 12°/km, a strain rate of 10^{-15}/sec, and a dry olivine flow law), and the brittle extension of the lithosphere will involve the upper crust only. However if the crust is thin, the lithospheric strength will be dominated by the upper mantle, and this layer should control the style of deformation.

The graben model of rift formation [3,4] predicts that the rift width w will be given by $w=k\pi\alpha/4$, where $1<k<2$, $\alpha=(EL^3/3g\Delta\rho(1-\nu^2))^{1/4}$, E is Young's modulus, L is the elastic layer thickness, g is gravitational acceleration, $\Delta\rho$ is the difference in density between the layers above and below the faulted layer, and ν is Poisson's ratio. For rifting of a crustal layer alone, $\Delta\rho$ is just the mantle density ρ_m, and w will be less than about 20 km [5], unless an unreasonably thick lithosphere is assumed (e.g. [6]). For a layer at depth with a strengthless layer above it $\Delta\rho=\rho_m-\rho_c$, where ρ_c is the crustal density. We can define an effective density contrast as $\Delta\rho=\rho_m-\beta\rho_c$, where $0<\beta<1$ is the degree of decoupling between displacements at the moho and the surface due to flow in the ductile lower crustal layer. Assuming $E=1.25\times10^{11}$ Pa, $\nu=.25$, $\beta=0.5$, $\rho_c=2.7$ Mg/m^3, $\rho_m=3.2$ Mg/m^3, and L=20 km gives the width for a graben formed in the mantle brittle zone of between 50 and 100 km. In addition, it can be shown using energy considerations [4] that surface graben depths in excess of 2 km are possible using these parameters. These results agree well with the observed dimensions of Venusian rifts, which typically have widths on the order of 75-100 km and depths of up to 2.5 km.

A layered rheology also offers an explanation for the spacing of probable faults associated with the rifts observed on Earth-based radar images. Subparallel bands of high radar backscatter in Beta Regio have been interpreted as fault scarps [7]. These features are superimposed on the rift zone, and have a characteristic spacing of roughly 20 km. Whereas the mantle layer controls the width of the rift, the thin surface layer should respond to its tensional and flexural deformation by

faulting at a characteristic spacing determined primarily by the thickness of the crustal strong layer. For reasonable assumptions about the rheology of the crust, a variety of models predict spacing of features in agreement with observations.

The graben model places constraints on crustal thickness and thermal gradient. The crust must be less than about 20 km thick (assuming a dry diabase flow law) to allow the existence of a mantle brittle zone. In addition, in order for there to be a ductile layer above this zone the crust must be at least about 5 km thick. The requirement that the mantle elastic layer thickness be $\geq$20 km implies a thermal gradient on the order of 15°/km or less. Lower thermal gradients will accommodate somewhat thicker crusts, as the brittle-ductile transition will be deeper and thus the crust can be thicker without eliminating the mantle brittle zone.

Folding: The mountain ranges of Ishtar Terra are characterized by a series of bands of greater and lesser radar backscatter [8]. This banded terrain has been interpreted to be the result of compressional folding of a mechanically strong surface layer; the 15-20 km spacing of the bands implies a 2-10 km thickness of this layer for a variety of elastic or viscous models [5]. These thicknesses are in agreement with those obtained from calculations based on laboratory measurements of crustal rocks. However the models require stresses of several hundred MPa, far in excess of the yield strength of rocks, unless a large number of shear-decoupled layers are postulated.

Applying a folding model utilizing a frictional-ductile strength envelope [9] to Venus, we find that the critical stress required for this more realistic rheology is less than 100 MPa, less than the yield strength under crustal conditions. A thermal gradient of about 15°/km is required by the observed spacing. These are upper bounds, as a rounding of the sharp brittle-ductile transition [10] will result in lower critical stresses and shorter wavelengths for a given thermal gradient. This model also implies that the crust is relatively thick in this region (with no zone of upper mantle strength), in agreement with thermal and gravity considerations [11,12].

Conclusions: Lithospheric strength envelopes provide a framework with which to explain the formation of both the large highland rifts and the narrow bands of Ishtar Terra. A simple graben model in which a zone of significant mantle strength is separated from a thin brittle surface layer by a ductile lower crust is consistent with the observed widths and depths for the equatorial rifts, as well as the spacing of associated normal faults. The Ishtar banded terrain is likely due to compressional folding of a thick crust at stresses less than 100 MPa.

References: [1] Brace & Kohlstedt (1980) **JGR 85**, 6248; [2] Banerdt & Golombek (1985) **LPSC XVI**, 23; [3] Vening-Meinesz (1950) **Bull Inst R Colon Belge 21**, 539; [4] Bott (1976) **Tectonophys 36**, 77; [5] Solomon & Head (1984) **JGR 89**, 6885; [6] Schaber (1981) **GRL 9**, 499; [7] Campbell et.al. (1984) **Science 226**, 167; [8] Campbell et.al. (1983) **Science 221**, 644; [9] McAdoo & Sandwell (1985) **JGR 90**, 8563; [10] Kirby (1980) **JGR 85**, 6353; [11] Banerdt (1986) **JGR**, in press; [12] Morgan & Phillips (1983) **JGR 88**, 8305.